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THE SEASONAL VARIATION OF PERMANENT
CURRENTS WITHIN BAFFIN BAY

ROBERT L. CERES

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THE SEASONAL VARIATION OF PERMANENT
CURRENTS WITHIN BAFFIN BAY

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Robert L. Ceres

THE SEASONAL VARIATION OF PERMANENT
CURRENTS WITHIN BAFFIN BAY

by

Robert L. Ceres

Lieutenant, United States Navy

Submitted in partial fulfillment of
the requirements for the degree of

MASTER OF SCIENCE

United States Naval Postgraduate School
Monterey, California

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CERES, P.

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Robert L. Ceres

This work is accepted as fulfilling
the thesis requirements for the degree of
MASTER OF SCIENCE
from the
United States Naval Postgraduate School

ABSTRACT

The seasonal variation of permanent currents within Baffin Bay is investigated. A model based on physical reasoning is developed, whereby the speed of surface currents may be calculated during winter months, when observations are not available due to ice coverage. Verification is made where possible utilizing late fall, and early spring data. Limitations, applications, and variability of results are discussed. Two important applications are: first, to add to the oceanographic knowledge of the Baffin Bay area, and second, to provide an additional input to computer programs already in existence for forecasting the drift of sea ice within Baffin Bay.

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1. Introduction

Numerous oceanographic cruises have been made into Baffin Bay in recent years primarily by Canadian and United States oceanographic vessels. The purposes of these cruises are to extend man's knowledge of this region, and to explore its commercial and military potential. This research, and other traffic as well, is hindered in large part by the yearly cycle of winter ice growth which generally leaves Baffin Bay accessible to shipping only during the short summer season, July through September. Thus, while a great deal is known about Baffin Bay during these summer months, the winter oceanographic regime is almost totally unexplored. It is, then, the first intent of this paper, to consider one aspect of this winter regime, that is, the surface circulation pattern.

One important aspect of extending knowledge of this area is the problem of forecasting the drift of sea ice. The purpose and usefulness of such forecasts is well established; they are presently being made in the Baffin Bay area on a routine basis by hand calculation. Both the Fleet Weather Facility, Monterey, California, and the U. S. Naval Oceanographic Office, are in the process of developing computer programs to accomplish this task; and, more specifically, Knodle (1) in 1964 developed a computer program for forecasting the wind drift of sea ice. Verification of test data utilizing this program indicated that, if the additional input of permanent currents could be introduced into the program, the accuracy of results might be significantly improved. This input of

permanent currents would be in the form of vectors at grid points of any scale desired and would be added vectorially to the present output, which considers only wind effects. While this final step of programming was not completed in this paper, the basic data for its completion are developed and examined in some detail, together with a discussion of the variability of the currents and the limitations of the methods employed.

2. Background

The first extensive attempt to determine the circulation pattern in Baffin Bay was made in 1928; and, while much has been done since that time, the work of Kiilerich (1939) (2) on the Danish Godthaab expedition of 1928, still stands as the most authoritative account of the summertime surface currents that exist within the Bay. The results of this work, as modified by the more recent observations, is shown in figure 1. Exhaustive descriptive accounts of this region are available from any number of sources (3, 4); so only a few distinctive features which are pertinent to the problem at hand will be mentioned. Again, with reference to figure 1, it should be noted that Baffin Bay is largely an isolated body of water with limited exchange taking place through the narrow channels to the north, and across the Davis Strait to the south. A ridge there forms a definite sill, with a limiting depth of less than 730m. The continental shelf is narrow on both the Greenland and Canadian sides of the bay, with steep slopes falling off rapidly into the one large basin, whose depths are in excess of 2000m.

The circulation is cyclonic with one small gyral to the immediate north of Davis Strait and one large gyral centered over the Baffin Basin. Mean velocities are shown in cm/sec.

Various climatological atlases show that, from mid-November through late May, Baffin Bay is virtually ice-bound; and thus in the following section a model is presented to show how the surface current, at any given location within Baffin Bay, is diminished with the advent of total ice coverage.

SUMMERTIME SURFACE CIRCULATION PATTERN BAFFIN BAY

(VELOCITIES IN CM/SECOND)

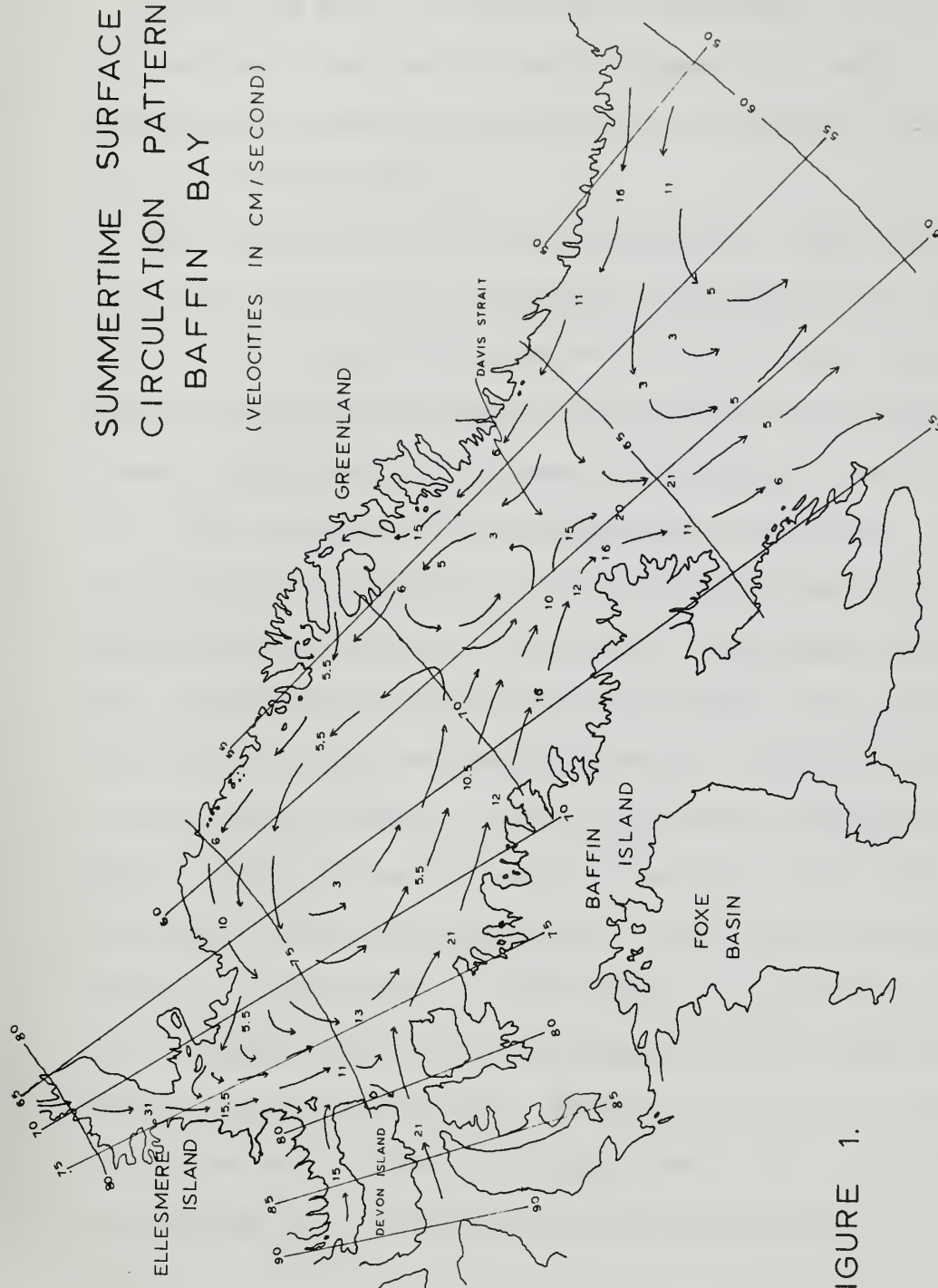


FIGURE 1.

3. Wintertime Regime

Due to the fact that no winter measurements have been made, this section is devoted to the development of a model, which, when applied, will describe the wintertime variation of the surface currents within Baffin Bay.

Two assumptions will be made initially: first, that the summertime circulation is primarily maintained by the wind, and further, that the contribution made to the surface circulation from the Arctic Basin and the North Atlantic is at a maximum during the summer, decreasing to a minimum by mid-winter. The latter part of this first assumption will be discussed in section 6. The former, is justified qualitatively on the basis of climatological wind data, and descriptive accounts of the area. For example, mean sea pressure charts for the summer months show Baffin Bay to be under the influence of a cold low pressure center. Further, the Polar Basin is under the influence of a cold high which contributes to steady easterly winds across the northern portion of Baffin Bay throughout the summer. The second assumption is that, once Baffin Bay becomes completely ice covered, no wind stress is transmitted through the ice. In other words, the contribution the wind makes towards maintaining the current is zero. Naturally, there will be some shifting of the ice due to tides, freezing effects, etc.; but in the mean the ice will be, for all practical purposes, stationary.

If the assumptions above hold, the only real forces acting to alter the basic currents are first, the force of friction between ice and water, and second, the frictional forces between the sea floor and the water. The latter is assumed to be relatively small and is not considered in this model. This is reasonable when one considers that the continental shelf is narrow, and that several investigators, including Kiilerich, found the deep-basin water to be near motionless. Further, if one assumes that the energy gained from currents entering the bay, is approximately equal to bottom and boundary losses, then the omission of these losses is further justified, at least qualitatively.

The fundamental winter relationship can now be seen: the energy of the currents is diminished at the rate at which energy is dissipated by frictional stress between ice and water. If this rate is known, together with the vertical distribution of velocity, it follows that the rate at which the current velocity decreases can be calculated. In order to find this rate, first let the stress be given by

$$\tau = k_w \rho_w u^2 \quad (1)$$

where k_w = coefficient of friction between ice and water,

u = surface current velocity, defined to be the current found a few meters beneath the sea surface, and

ρ_w = density of the water at the same level

The rate of work per unit area is τ_u , which is to be equated to a rate of energy dissipation per unit area:

$$\tau_u = -\frac{d}{dt} \frac{(\text{Kinetic Energy})}{\text{Area}} \text{ or,}$$

$$\tau_u = k_w \delta_w \dot{u} = -\frac{d}{dt} \left(\frac{1}{2} [\bar{u}_1^2 + \langle u_1'^2 \rangle] \rho_1 \Delta h + \frac{1}{2} [\bar{u}_2^2 + \langle u_2'^2 \rangle] \rho_2 \Delta z \right) \quad (2)$$

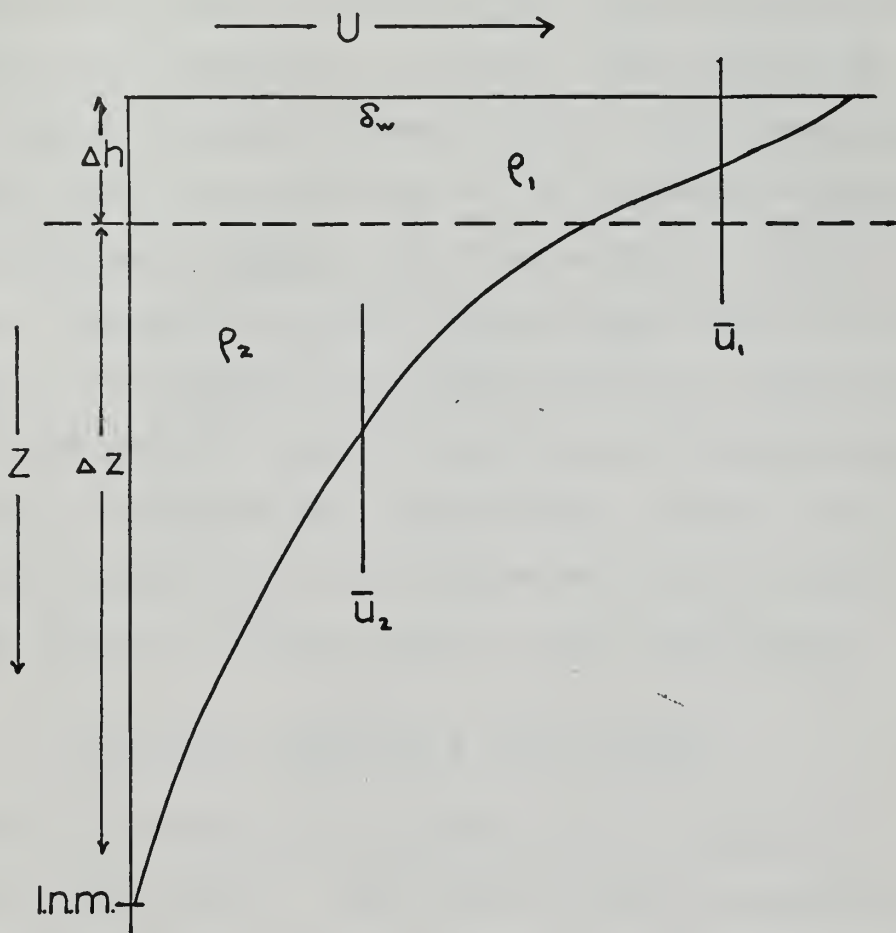
where \bar{u}_1 = mean current speed in the friction layer,

u_1' = the deviation from the mean \bar{u}_1 ,

ρ_1 = the mean density in the friction layer,

and, Δh = depth of the frictional layer.

The corresponding parameters in the layer from the depth of friction, to the level of no motion are \bar{u}_2 , ρ_2 , and Δz as shown in figure 2. Note that $\bar{u}^2 = \bar{u}^2 + \langle u'^2 \rangle$ is applicable in either layer.



Schematic representation of a
wind-driven velocity profile
showing quantities represented
in equation 2.

FIGURE 2

From the observations made of velocity profiles shown in Appendix A, two initial observations can be made. First, the profile can change radically within a short period of time, and within short distances; and, second, these changes take place throughout the entire column of water. While these profiles may contain many inaccuracies as will be discussed in section 6, the observations do suggest that a loss or gain of energy at the surface influences the entire column of water that is in motion. Thus, the assumption is made that the kinetic energy loss is distributed from the surface to the level of no motion nearly uniformly. This means that the total loss, that is, the right hand side of equation (2), may be expressed solely in terms of the loss within the friction layer, or more specifically

$$k_w \int_w \bar{u}^2 = - \frac{d}{dz} \left(\frac{C}{2} [\bar{u}_1^2 + \langle u_1'^2 \rangle] \rho_1 \Delta h \right) \quad (3)$$

where the constant C is dependent upon the level of no motion (LNM) and Δh . The level of no motion applicable to Baffin Bay is discussed in section 4; but, if for example, the LNM is found at 1000m, and $\Delta h = 50m$, then C would equal twenty etc. (or $C = LNM/\Delta h$). This doubtless is an oversimplification. This assumption, however, appears to fit the observations, and gives realistic results.

Thus, in equation (3), the total kinetic energy loss is expressed solely in terms of the loss that occurs within the friction layer. This layer will be examined in some detail in the

remainder of this section. If several basic relationships, taken from Shuleikin (5), relating to wind drift of sea ice are introduced, it will be shown that the thickness of the friction layer, Δh , may be expressed purely as a function of the surface current velocity, and further that this same expression may be used for the model under consideration. Thus, for a current flowing beneath stationary ice:

$$(a) \Delta h = \pi \sqrt{\mu / \delta_w \bar{\omega}},$$

$$(b) u = T \Delta h / \mu \pi \sqrt{2},$$

$$(c) u = AV / \sqrt{5 \sin \phi},$$

$$(d) T = k_a \delta_a V^2,$$

where:

T = tangential stress between air and ice,

μ = eddy viscosity coefficient,

δ_a = density of the air,

$\bar{\omega} = \omega \sin \phi$, where ω is the angular velocity of the earth's rotation, and ϕ = latitude,

k_a = coefficient of friction between air and water

A = the wind factor,

V = the wind velocity, and

$\Delta h, \delta_w, \mu$, are as previously defined.

From (b) and (c) is obtained $\mu = \frac{T \Delta h \sqrt{\sin \phi}}{\pi \sqrt{2} A V}$,

which when substituted into (a) yields

$$\Delta h = \frac{\pi T \sqrt{\sin \phi}}{\sqrt{2} \rho_w \sin \phi A V} = \frac{\pi (k_a \delta_a V^2) \sqrt{\sin \phi}}{\rho_w \omega \sin \phi \sqrt{2} A V} ,$$

Finally, solving (c) for V and substituting into the above,

$$\Delta h = N u, \quad \text{where} \quad N = \frac{\pi \delta_a k_a}{\sqrt{2} \rho_w \omega A^2} .$$

With sufficient accuracy $k_a = 2 \times 10^{-3}$, and on the basis of studies by several authors, $A = 1.27 \times 10^{-2}$. Substituting these, and other known values, into the above expression yields

$$\Delta h = 473 u \quad (4)$$

where Δh is in meters if u is in meters/sec.

In the above development the tangential stress T operating on the sea surface was caused by wind. Shuleikin concluded, however, that with ice coverage the frictional layer would also be present, but would be generated by the surface friction between ice and water. Further, he concludes that the numerical value of the coefficient N will not vary perceptibly because "it is known that the turbulent regime affecting the value Δh becomes established under the influence of any velocity of the surface current u, independent of whether this velocity is generated by the tangential stress of wind origin, or by the stress caused by the friction of ice on the water". Thus, in this model equation (4) will be used to relate u and Δh .

With the magnitude of the currents that we are dealing with, it is easily seen that this depth of friction will rarely exceed fifty or sixty meters; and within this thin upper layer it can be assumed with a good degree of accuracy that $\delta_w \doteq \rho_1$, where δ_w is the density of the water a few meters beneath the surface, and ρ_1 is the mean density within the friction layer.

The numerical value of the coefficient of friction, k_w , has been investigated in some detail by several authors, and in particular by Brown and Crary (6), Fukutomi (7), and Shuleikin (5). The values varied with location and with the age of the ice. Wittman and MacDowell (8) modified Shuleikin's coefficients for use specifically in Baffin Bay, and Knodle (1) concluded after further investigation, that a mean value of .013 was the best approximation. This value of k_w then will be used for this model.

Now, from equation (4) and the density approximation above, equation (3) takes the form,

$$k_w u^3 = -\frac{1}{\rho_1} \left(\frac{\rho_1}{2} u^2 + \langle u'^2 \rangle \right) 4\pi \eta u \quad (5)$$

It will be shown in the following paragraphs that the expression $[\bar{u}^2 + \langle u'^2 \rangle]$ may be treated, with some approximation, solely as a function of the surface velocity, u , if it is assumed that the shape of the velocity profile, once established, does not alter significantly. While the wind-driven profiles given in appendix A show a great variety of shape, it should be noted that the greatest

variability occurs when the currents are weak and in shallow water. Profiles associated with stronger currents in deep water (i.e., from 5 to 10 cm/sec. at the surface) show a definite uniformity of shape. It is argued here that most of the profile variability comes about as the result of non-uniform influences, such as wind gradients at the surface, summer run-off, inflowing currents, tides, or internal waves, and that these disturbing influences are either totally absent or at a minimum during the winter season.

Considering the above, we will now construct a model for the current velocity profile. First, it is assumed that during the winter, the frictional component of the velocity (V_f) is analogous to that given by Ekman; then

$$(e) \quad V_f(z) = u_{\Delta h} e^{-\pi z / \Delta h}$$

where, $u_{\Delta h}$ = velocity found at the depth of friction, and

$V_f(z)$ = frictional component of the velocity found at depth z within the friction layer.

Further, it is assumed that, due to the proximity of land within Baffin Bay, there is no rotation of this frictional component with depth. It follows that:

$$(f) \quad u_z = u_{\Delta h} - V_f(z) ,$$

where u_z = the magnitude of the current found at depth z within the friction layer. If, for example, at the time the ice becomes stationary, the depth of friction is found at 50m, then, by equation (4), $u_{\Delta h}$ may be calculated and a velocity profile plotted (see figure 3).

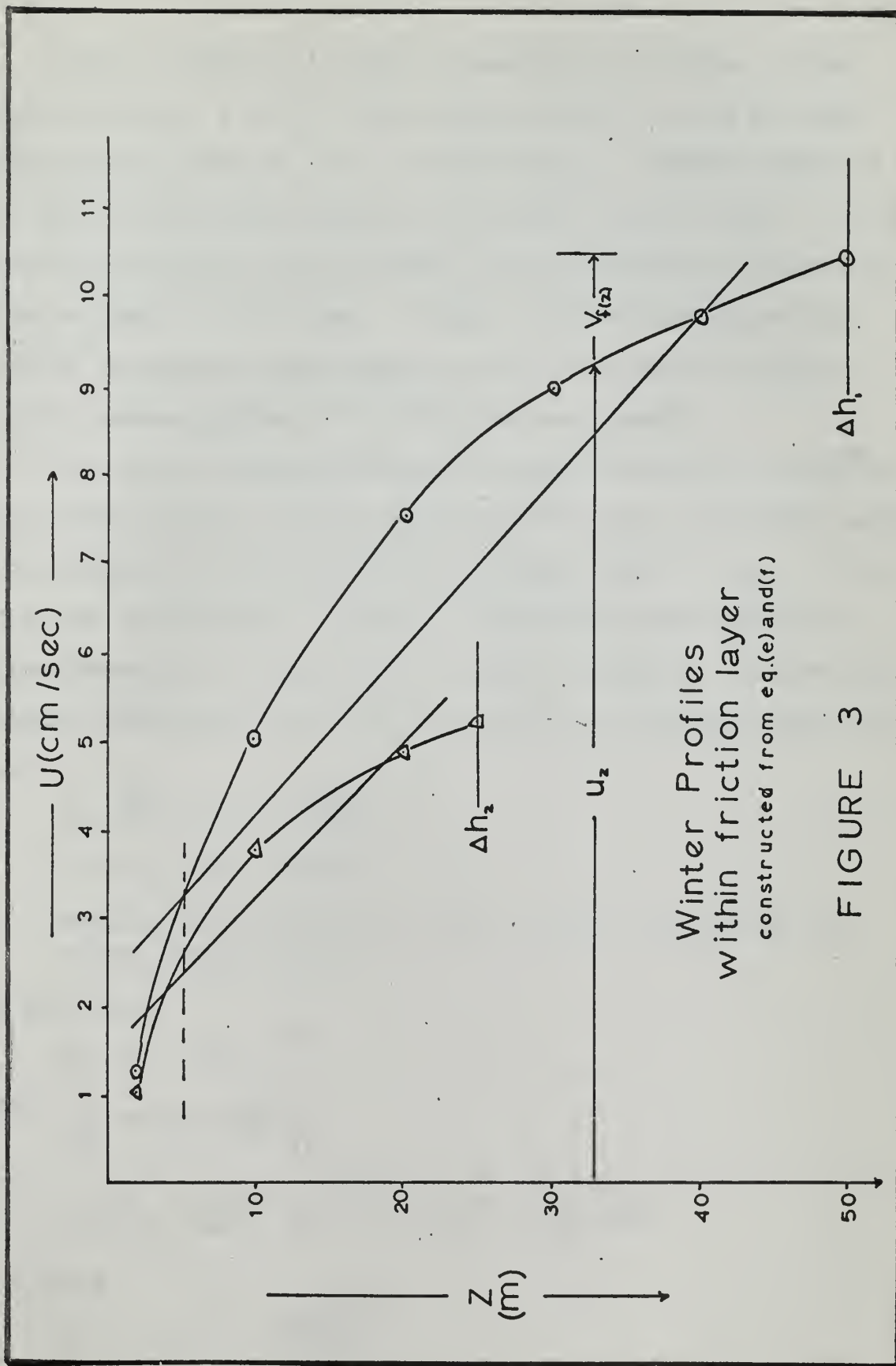


FIGURE 3

Several comments are needed concerning this profile. First, the velocity at $z = \Delta h$ is the proper velocity to give the relation between u and Δh , i.e., $\Delta h = 473 u_{\Delta h}$. Second, equations (e) and (f) are being utilized as the winter profile model since no direct and detailed profile studies for flow beneath stationary ice were available to the author. Finally, the introduction of this profile necessitates redefining u , the surface current velocity, as will become apparent in the following paragraphs.

To find the relation between the surface velocity u , and $\overline{u^2}$ in the friction layer, several velocity profiles were constructed utilizing equations (e), (f), and (4) for assumed values of $u_{\Delta h}$, from 12 cm/sec to 4 cm/sec. For ease of computation each profile was approximated by a straight line as shown in figure 3. The velocity gradient with depth could then be treated as a constant ($du/dz = C$), and

$$(g) \quad \overline{u} = u_5 + C \frac{\Delta h}{2}$$

$$(h) \quad u_z = u_5 + C z$$

where u_5 is the velocity at depth 5 meters, defined as the surface current velocity.

By definition

$$(i) \quad u' = u_z - \overline{u}.$$

Thus
$$u' = C \left(z - \frac{\Delta h}{2} \right),$$

or

$$\langle u'^2 \rangle = \frac{1}{\Delta h} C^2 \int_0^{\Delta h} \left(z^2 + \Delta h z + \frac{\Delta h^2}{4} \right) dz$$

and hence

$$(j) \quad \langle u'^2 \rangle = \frac{\Delta h^2 C^2}{12}$$

By utilizing equations (g) and (j) to calculate the quantities $\langle u'^2 \rangle$ and \bar{u} for each velocity profile approximation, two important conclusions were reached: first, that the quantity $[\bar{u}^2 + \langle u'^2 \rangle]$ could be approximated quite accurately by $[\bar{u}^2 + \langle u'^2 \rangle] = 1.1 \bar{u}^2$, and secondly, that the same expression can be described in terms of the surface current velocity u , again without significant loss of accuracy

$$[\bar{u}^2 + \langle u'^2 \rangle] = 1.1 \bar{u}^2 = k_1^2 u_s^2, \quad (5a)$$

where the constant k_1 equals 2.32. If equation (5) is modified in accordance with equation (5a), then

$$k_w u_s^3 = -\frac{d}{dt} \left(\frac{c}{2} [k_1^2 u_s^2] 473 u_{\Delta h} \right). \quad (6)$$

With the rapid decrease in current velocity near the ice surface associated with the winter profile of figure 3, it becomes apparent that u_s no longer adequately defines the velocity within the friction layer, but rather that $u_{\Delta h}$ is the representative current velocity. Utilizing the same winter profiles as in the above development of \bar{u}^2 , it was found that the approximation $u_{\Delta h} = 3.18 u_s$ was quite accurate, providing the range of velocity profiles chosen is not too large. That is, within the range utilized, 4-12 cm/sec for $u_{\Delta h}$, the accuracy is quite good, and would be the range of values expected for the major portion of Baffin Bay. Hence, if we define $u = u_{\Delta h}$, and substitute the above approximation into equation 6, then its solution is

$$u(t) = u(t_0) e^{-\frac{k_w}{3k_2}(t-t_0)} \quad (7)$$

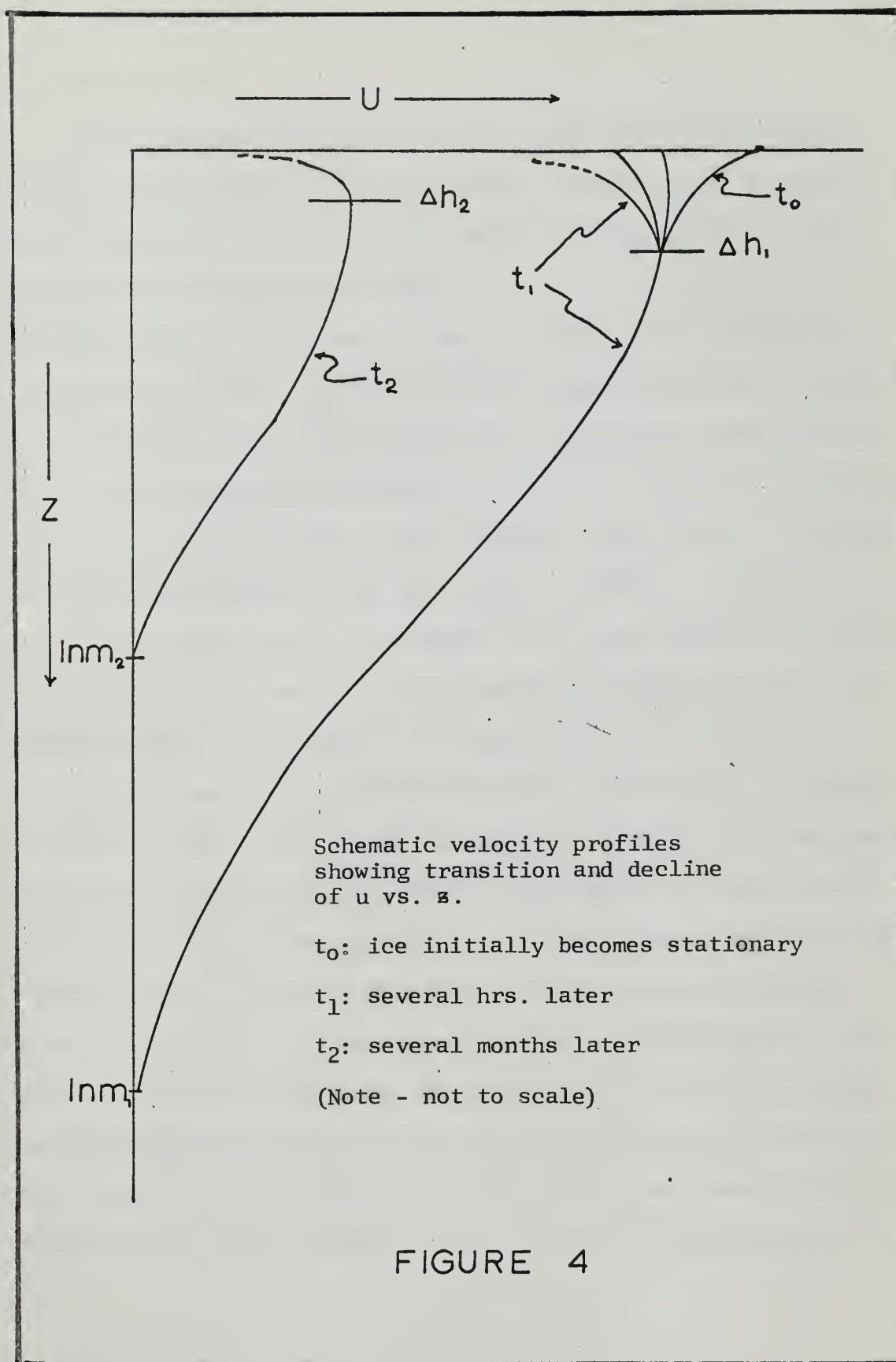
where

$$k_2 = \frac{473}{2} (3.18)^2 c k_1^2$$

It is seen then, that at the time the ice becomes stationary the current velocity u of equation (7) represents the current found at the depth of friction, and that this depth will decrease as the winter season progresses approaching the surface by the time of spring ice break-up (see figure 4). Thus, u at the depth of friction is applicable only during the time the ice is stationary, while u several meters beneath the sea surface (i.e., at 5m) is applicable for the remainder of the year.

Thus, equation (7), shows the time variation of the current at depth Δh , where $u(t_0)$ represents the current at the initial time of ice coverage and is known if the assumption is made that the current velocities of figure 1 are representative of $u_{\Delta h}$ at the time the ice becomes stationary. Hence, we may calculate the current at any time $u(t)$ between initial ice coverage and the time of break-up of the ice. The application of this equation is discussed in the following section, and a curve showing the seasonal variation of the surface current is developed utilizing this equation.

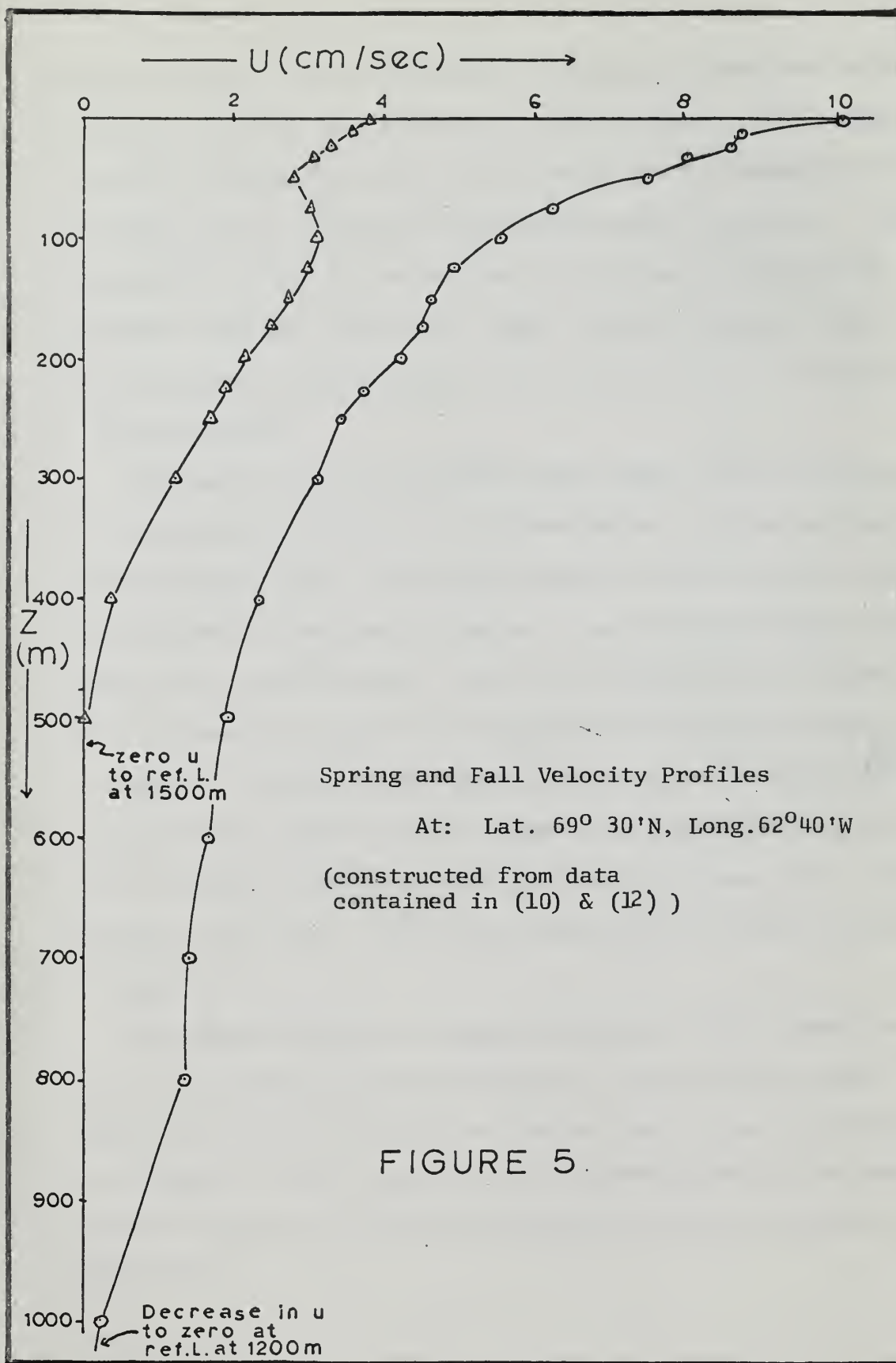
The transition from the summer to the winter profile, is believed to take place within a short time after the ice becomes stationary. This is analogous, for example, to tidal currents near bottom, as given by Defant (9). In this transition period the K. E. loss is confined almost entirely to the friction layer, and the reduction of K. E. with depth does not occur until after the winter profile becomes established. This transition period is shown schematically in figure 4, together with a profile, say several months later.



4. Application

In this section the application of equation (7) developed in the previous section will be discussed, together with a qualitative treatment of the spring and summer current build-up, thus completing the mean yearly cycle.

Two facets of equation (7) must be dealt with prior to its application. First, the time the ice becomes stationary and the time break-up of the ice occurs must be determined; and, second, the value of C , discussed briefly in section 3, needs to be calculated. The latter will be treated first. Recall that $C = LNM/\Delta h$ and that the assumption has been made that kinetic energy is dissipated from the surface to the level of no motion nearly uniformly. This implies that the level of no motion descends with time. By equation (4) it is seen that the depth of friction Δh is a function of the surface current velocity u and thus is likewise a function of time (see figure 4). Hence, a profile study was undertaken, one purpose of which was to investigate the variation of these quantities (see appendix A). From it, some approximations could be made. From the fall data, it was estimated that the level of no motion in deep water occurred at approximately 1000m and, as previously discussed, $\Delta h = 50m$. Only a very few spring soundings were available to the author, and only one in deep water. The velocity profile constructed from them is as shown in figure 5, together with a fall velocity profile taken at approximately the



same location. While this indeed is sketchy information, both the spring and the fall profiles, in this instance, lend support to earlier assumptions, and in particular to the assumption that an energy loss is dissipated with depth nearly uniformly. From this profile it was estimated that at the time of break-up of the ice $LNM = 500m$ and $\Delta h = 20m$. Thus, C varies from the fall to the spring only from 20 to 25; a mean value of 22.5 is chosen for C in equation (7).

The span of time during which this model would be applicable was determined on the basis of climatological data to be from October through April. The limitations of utilizing these dates are discussed in section 5; however, they should be approximately correct for an average year. Now all the quantities of equation (7) are known, and $u(t)$ may be calculated utilizing the velocities of figure 1 as $u(t_0)$. The expression from equation (7), $e^{-\frac{k_w}{3K_2}(t-t_0)}$, represents the fraction of the summer time current velocity and may be plotted versus time with $t_0 = 0$ on 1 October and t (in seconds) up to 1 May. The plot is shown by the solid line portion of figure 6.

The assumed spring and summer build-up of the current is shown by the dashed portion of figure 6. This build-up was handled in a purely qualitative fashion due to time limitations. Now, figures 1 and 6 together permit determination of the mean current velocity for any time during the year or location within Baffin Bay.

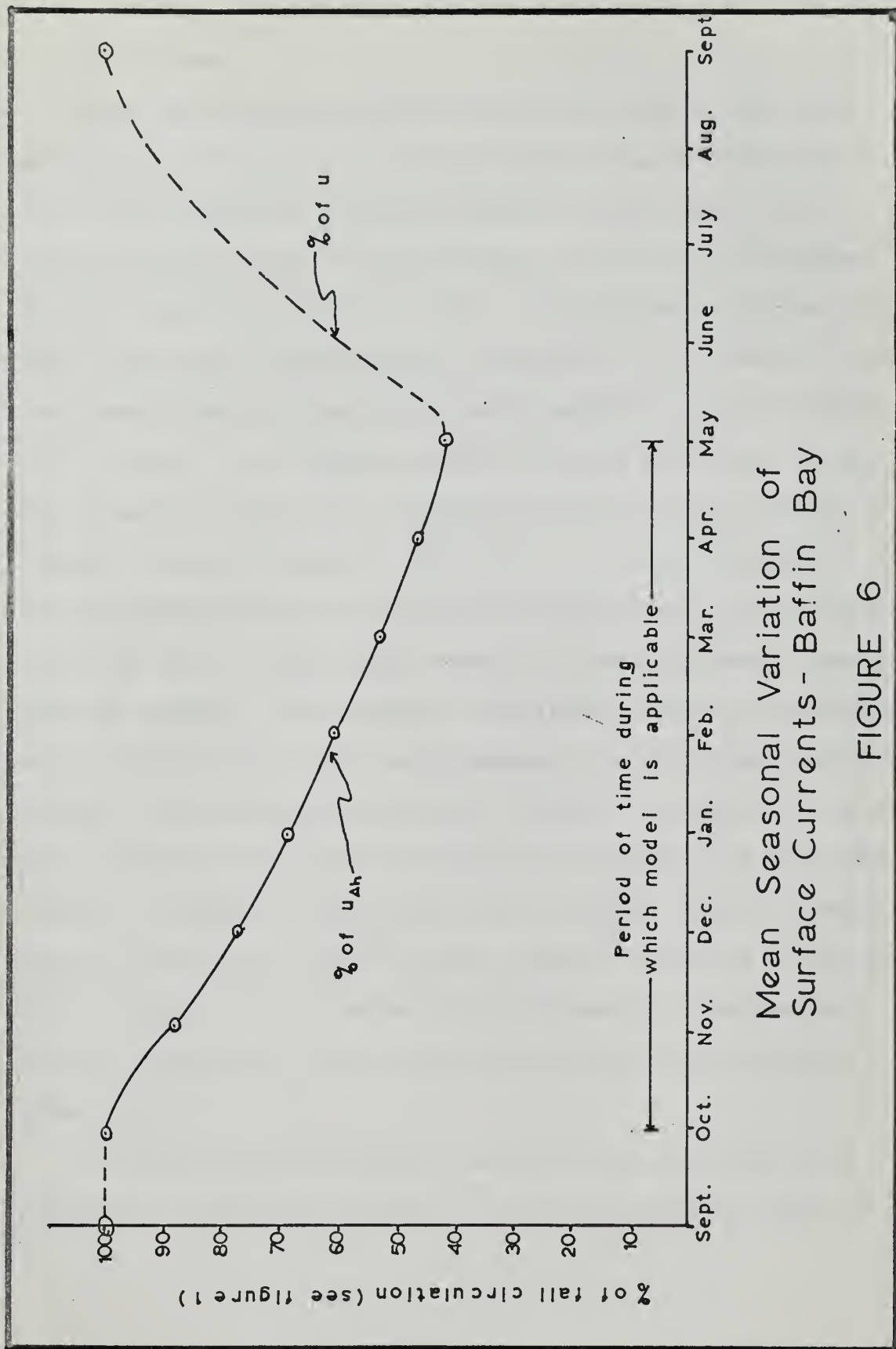


FIGURE 6

5. Limitations

There are three major limitations to the winter-time model developed in the preceding sections which require discussion. First, the assumption was made, based on climatological data, that the ice cover became approximately stationary on 1 October, and that break-up occurred on 1 May. This statement implies that these times occur simultaneously throughout the Bay, when in fact the freezing and break-up occur over a period of time with varying locations. An attempt was made to divide Baffin Bay in four approximately equal areas and develop separate curves similar to figure 6 for each quadrant. This trial was abandoned, however, as it introduced more uncertainties and complications into the problem, and these disadvantages seemed to outweigh whatever benefits might be derived. For example, a declining current in one quadrant would certainly effect the next quadrant into which the current was flowing. No ready solution to this problem is proposed; it is felt that whatever modification is introduced would lead to only small changes in figure 6. The use of climatological data also implies that any one year is like the next, and this obviously is not the case. Figure 6 could, however, be quite easily reconstructed, based on information from local reconnaissance flights for any given year.

The second major limitation concerns the variation in the summer time build-up of currents. Evidence indicates that this

variation in some years might be large. For example, in 1964 residual ice was observed in Melville Bugt throughout the summer, because the mean pressure patterns normally associated with this season were displaced enough from their normal positions that the winds, which normally would advect ice out of Baffin Bay and contribute to the current build-up, were simply not present. During such an anomolous year, the model would obviously not work. However, this unusual situation is a rarity.

The final limitation concerns the sum of the approximations and assumptions made in the development of the model itself. While these assumptions and approximations have all been discussed individually and justified, the cumulative effects could be additive and result in serious error. Particularly in the assumption of the shape of winter velocity profile and in the treatment of $C = LNM/\Delta h$ could large errors arise. However, in the absence of winter data, it is not possible to assign a number to this cumulative error. In section 7, verification of a forecast utilizing this model is quite good. This in itself is not proof that all the approximations and assumptions are valid, but it does at least lend support.

6. Current Profile Study

In order to verify the assumptions made in earlier sections, as well as to verify figures 1 and 6, an attempt was made to obtain the published results from the latest fall and earliest spring oceanographic cruises that had been made into Baffin Bay. These data were received from the Bedford Institute of Oceanography, Dartmouth, N.S., via the Canadian Oceanographic Data Center (10, 11). Utilizing the standard techniques of computing and summing dynamic height anomalies, a number of velocity profiles was constructed. These profiles, together with their locations, are shown in appendix A. While, in large part, this study failed to yield the verification that was hoped for, some interesting observations can be made. The remainder of this section is devoted to a discussion of what was hoped to be gained by this study, whether success or failure was met, and whenever possible why.

First, it was hoped that the shape of the velocity profiles, at least for the fall data, would fit the preconceived notion of a wind-driven velocity profile (see figure 2). While in a number of instances, particularly in deep water, this was the case, a great variety of shapes resulted. While causes for variability were mentioned briefly in section 3 (tides, internal waves, etc.), it is believed by the author that a major limitation of the dynamic method of computing currents is that oceanographic station data cannot be taken simultaneously (that is, with only one ship). For the

profiles shown in appendix A, a lapse time of from two to six hours occurred between adjacent stations; if reorientation of the dynamic topography were to take place during this interval, distortion in the magnitude of the current at a given depth would occur.

It was hoped, secondly, that the level of no motion could be firmly established for the spring and the fall, as various investigators, and particularly Kiilerich, had stated that this level will vary from season to season, perhaps even irregularly from place to place. While some substantiation of the figures used in this model were gained from this study (see figure 5), it can be noted from appendix A, that this level cannot be fixed with confidence. Here also, one must be careful to distinguish between the reference level and the level of no motion. For this study, the reference level was chosen in every instance at the greatest depth that the adjacent soundings would permit. However, in some instances the soundings did not approach the bottom; and, hence, the profiles give only a velocity that is relative to whatever current exists at this reference level. This of course means also, that the true level of no motion could be at quite a different depth from that implied by the profiles.

One of the initial assumptions made was that exchange of water between Baffin Bay and the bodies of water to the North and South was at a maximum during the summer, decreasing to a minimum by mid-winter. As can be seen from appendix A, only the exchange across Davis Strait

could be examined. While the assumption made cannot be validated on the basis of this study, it is interesting to note that, if the assumed level of no motion is correct, outflow greatly exceeded inflow in both seasons with no inflow occurring with the fall data, and very little with the spring data. From previous calculations of current flow across Davis Strait, Dunbar (4) had stated that the volume transport is approximately in the ratio of 2 to 1 with outflow being twice as great. He concluded therefore, that Baffin Bay received its waters in approximately equal quantities from the north and the south, but failed to consider runoff, which on the basis of this study would appear to make an important contribution. Hence, while the ratio of 2 to 1 would appear to be an underestimation, if, in fact, water does enter Baffin Bay in approximately equal quantities from the North and South, one could draw the tentative conclusion that the amount of water entering Baffin Bay is small during all seasons.

On the basis of the fall data, an effort was made to validate the current velocities given in figure 1. Of the twenty-five fall velocity profiles plotted in appendix A, fourteen, or 56%, were in good agreement with figure 1, that is, within ± 3 cm/sec at the surface. One further limitation of the dynamic anomaly technique is that it gives only that component of the current which is normal to a line drawn between adjacent stations. This surely accounts in large part for the variation from figure 1 in several of the profiles.

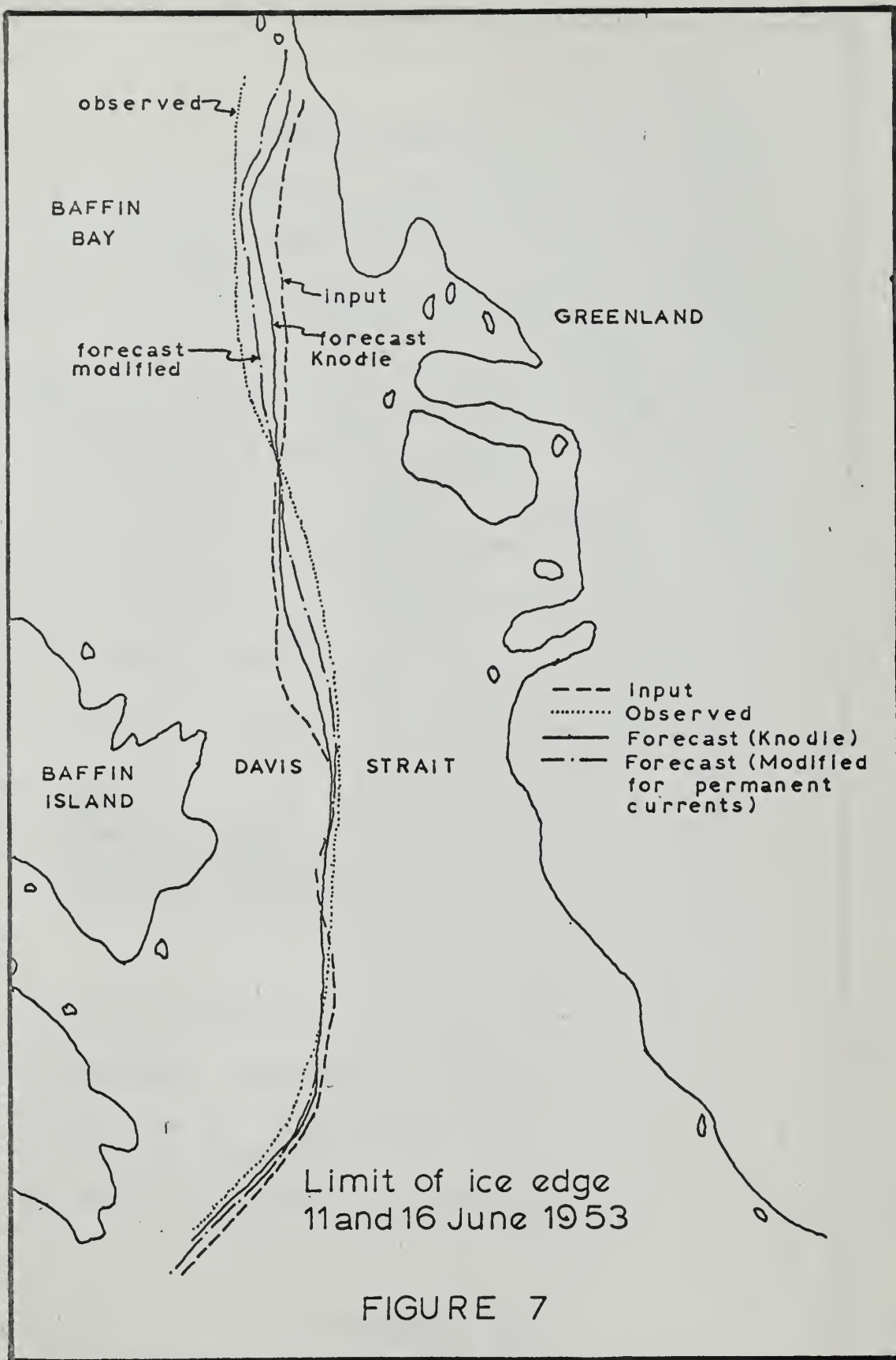
The remainder, however, and some are the exact opposite of what one would expect, can only be explained in terms of the limitations of the methods employed outlined in previous paragraphs. In conclusion, there was only fair agreement between the observations and figure 1, with no consistent error being observed that would indicate that it should be changed.

A verification of figure 6 was hoped to be accomplished on the basis of the spring data. These data, however, due to ice coverage were restricted primarily to Labrador waters, with only eight soundings taken within Baffin Bay. Of these eight, all but one showed a definite decrease in the velocity as the model predicts. The verification presented in the next section is another test of the model, and of figure 6.

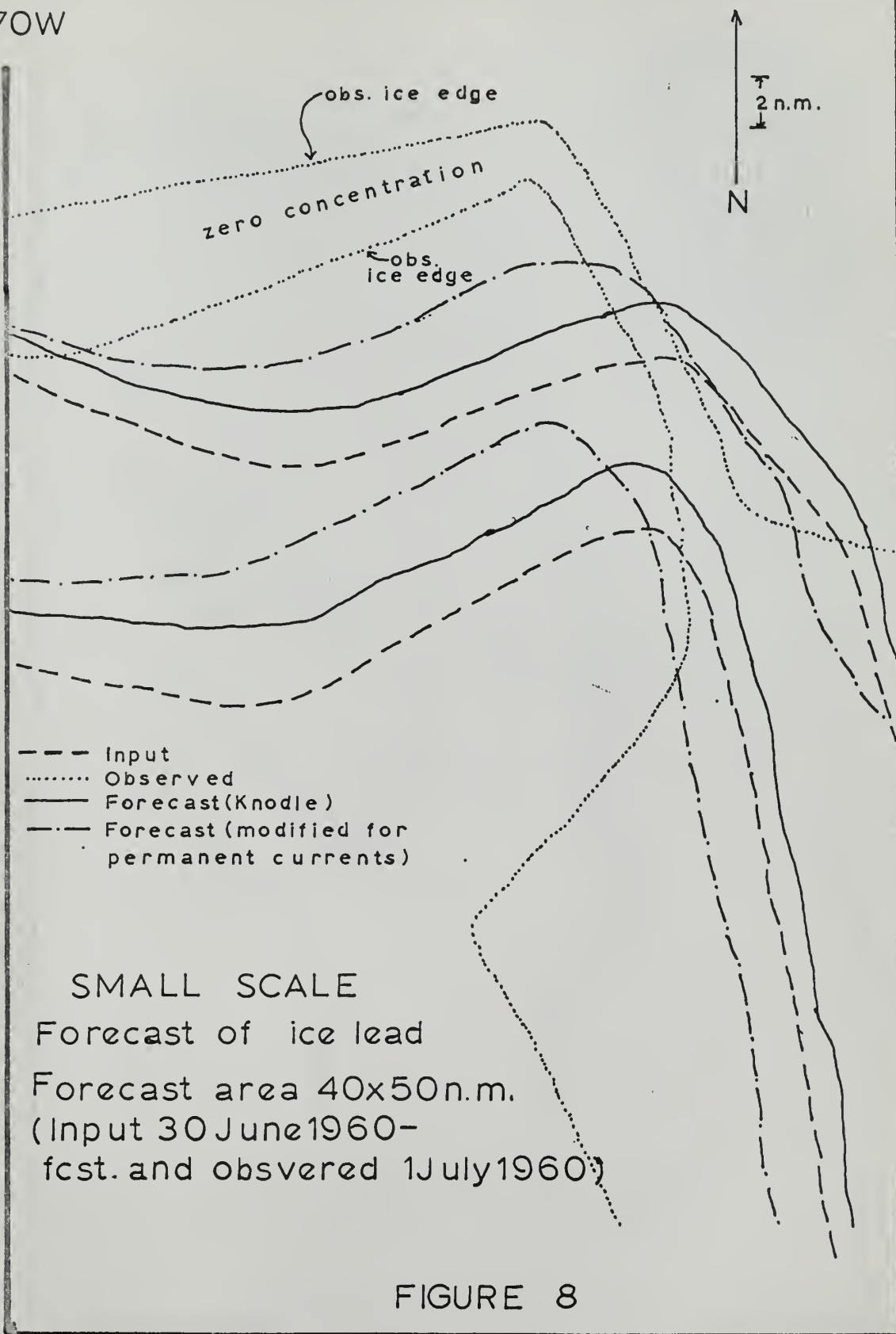
7. Verification

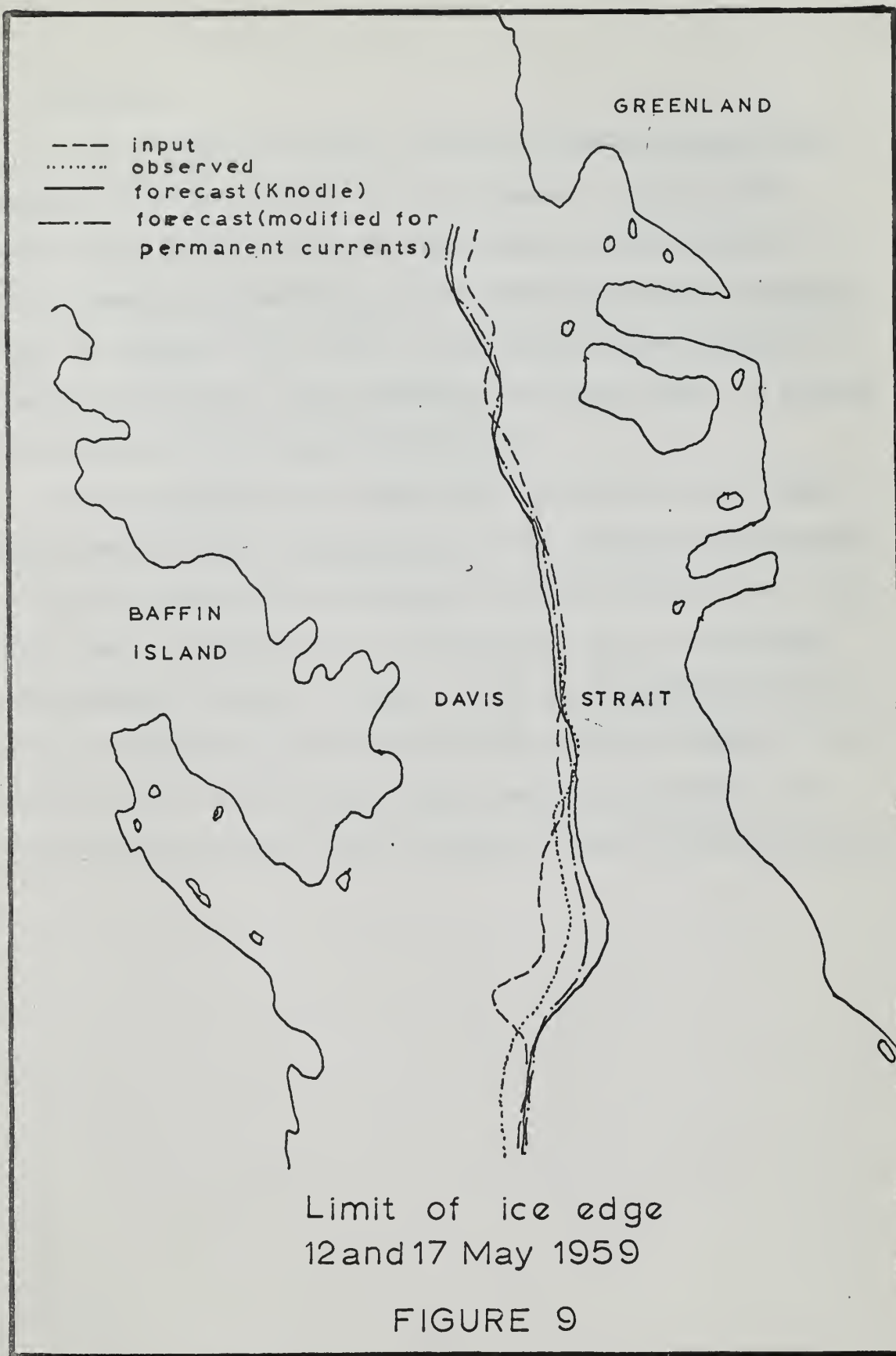
In Section 1, it was stated that one of the primary reasons for investigating the seasonal variation of the permanent currents within Baffin Bay was that the results of the study could be used as an additional input to Knodle's computer program for forecasting the wind drift of ice, developed specifically for Baffin Bay. In this section, the results of Knodle's three verification runs are shown in figures 7, 8, and 9, together with a modified forecast which includes the effects of permanent currents. This latter forecast was arrived at by adding to the wind-drift calculations of Knodle's, hand calculated, advection of the ice edge by currents in accordance with figures 1 and 6. While this is a laborious and time consuming task, the results were gratifying; and the job would be a simple one for the computer.

As can be seen, there was indeed a significant improvement over Knodle's results in each of the three cases. One might conclude, however, that the magnitude of the currents was too small, or, in effect, that figure 6 overestimates the winter seasonal decline of the current. It is possible that melting might account for a large part of the remaining error. It is concluded on the basis of the foregoing that, while figures 1 and 6 may be altered in the light of later findings, their use in forecasting the drift of sea ice within Baffin Bay tends to improve the drift forecasts.



+75 N
70W





8. Conclusion

In the preceding sections, a model has been developed which predicts the winter decline of the permanent currents within Baffin Bay. In the absence of winter data, it was necessary to make a number of assumptions, on the basis of physical reasoning, about the nature of the flow. The spring and summer build-up was handled qualitatively thus completing the yearly cycle. A limited verification of the results was obtained.

It is concluded that, despite the limitations of the model, steps have been taken toward adding to the oceanographic knowledge of the Baffin Bay area and providing the basic material for a computer input to forecasting the component of sea ice associated with permanent currents. Finally, it is recommended that the results of this paper be combined with the results of Knodle. It is felt that there would result a fully operational program, far faster and more accurate than is presently done by hand calculation.

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APPENDIX A

VELOCITY PROFILES

On the following pages are shown the velocity profiles as constructed from the data contained in (T0) and (11), together with the locations for the fall profiles (figure A. Sept. 17 - 24, 1960) and the spring profiles (figure B. June 1 to 11, 1963).

The computations were made in accordance with the Sandstrom-Holland-Hansen method of computing currents. It is assumed that the reader has some familiarity with the technique, so that only a brief outline will be required here.

1. Dynamic height anomalies were computed for the standard levels and summed upwards from the reference level (R.L.)

to the surface in accordance with: $\Delta D = \int_{p_1}^{p_2} \delta \Delta p$

with the specific volume anomaly δ . The R.L. was picked in all cases at the maximum depth that adjacent soundings would allow.

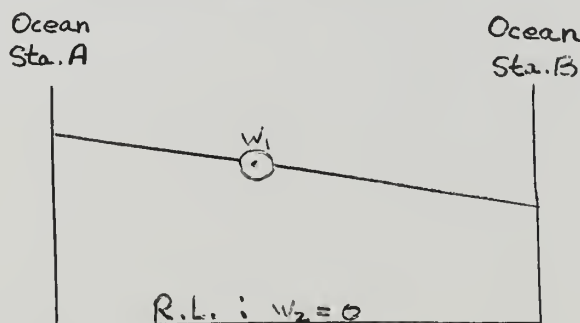
2. Relative currents for each level were computed by the formula:

$$w_1 - w_2 = \frac{10}{2f\omega \sin \phi} (\Delta D_A - \Delta D_B)$$

where

w_1 = current found at a given level

w_2 = current found at the reference level
assumed equal to zero



The results of these computations, as stated, are shown in the following pages together with other pertinent data such as the depth of the water, the time the sounding was taken, etc.



FIGURE A

Fall Data: 17 to 22 Sept. 1960

Location of Ocean Stations

57 through 89. Direction of surface
current between stations as indicated

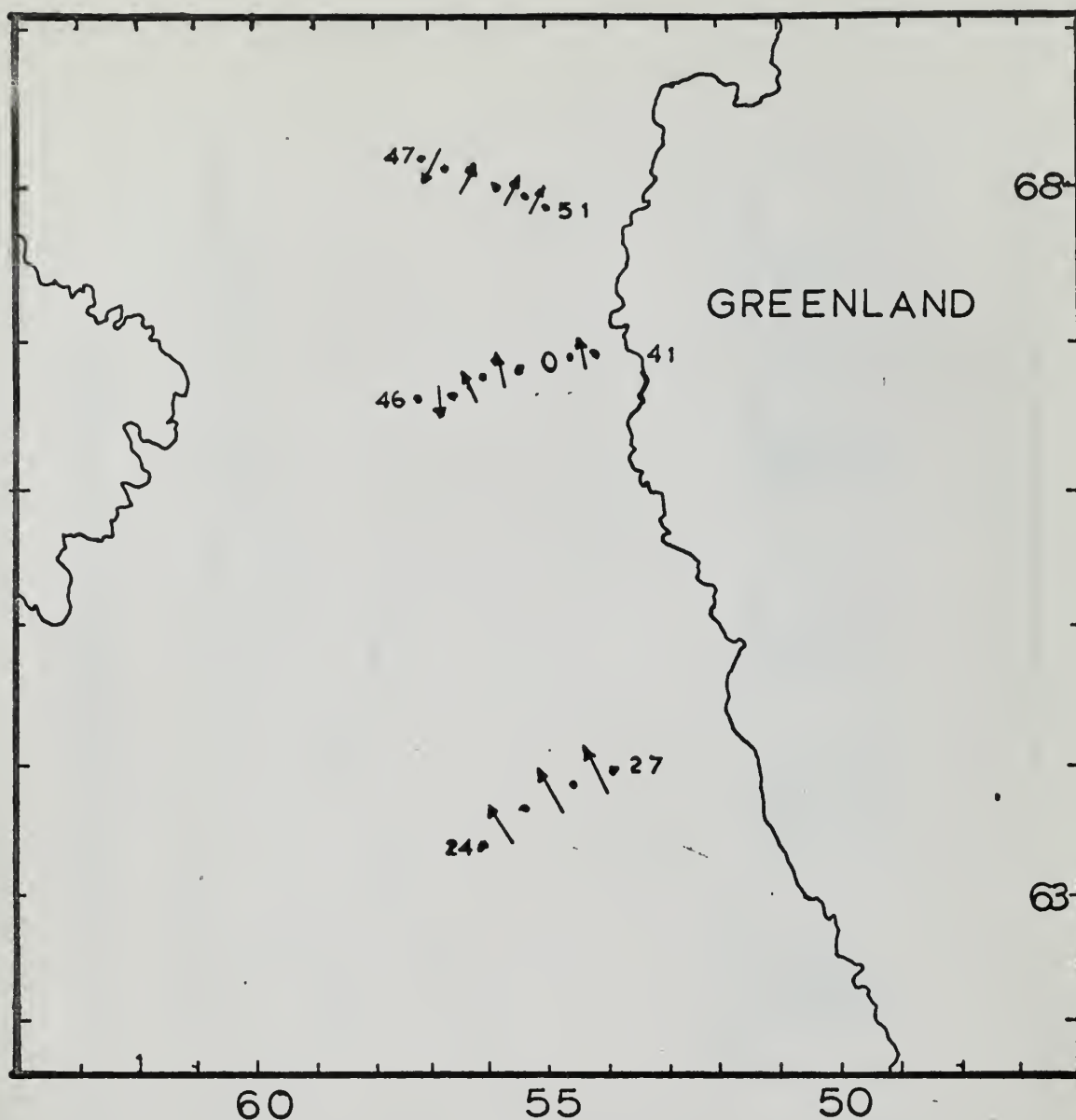
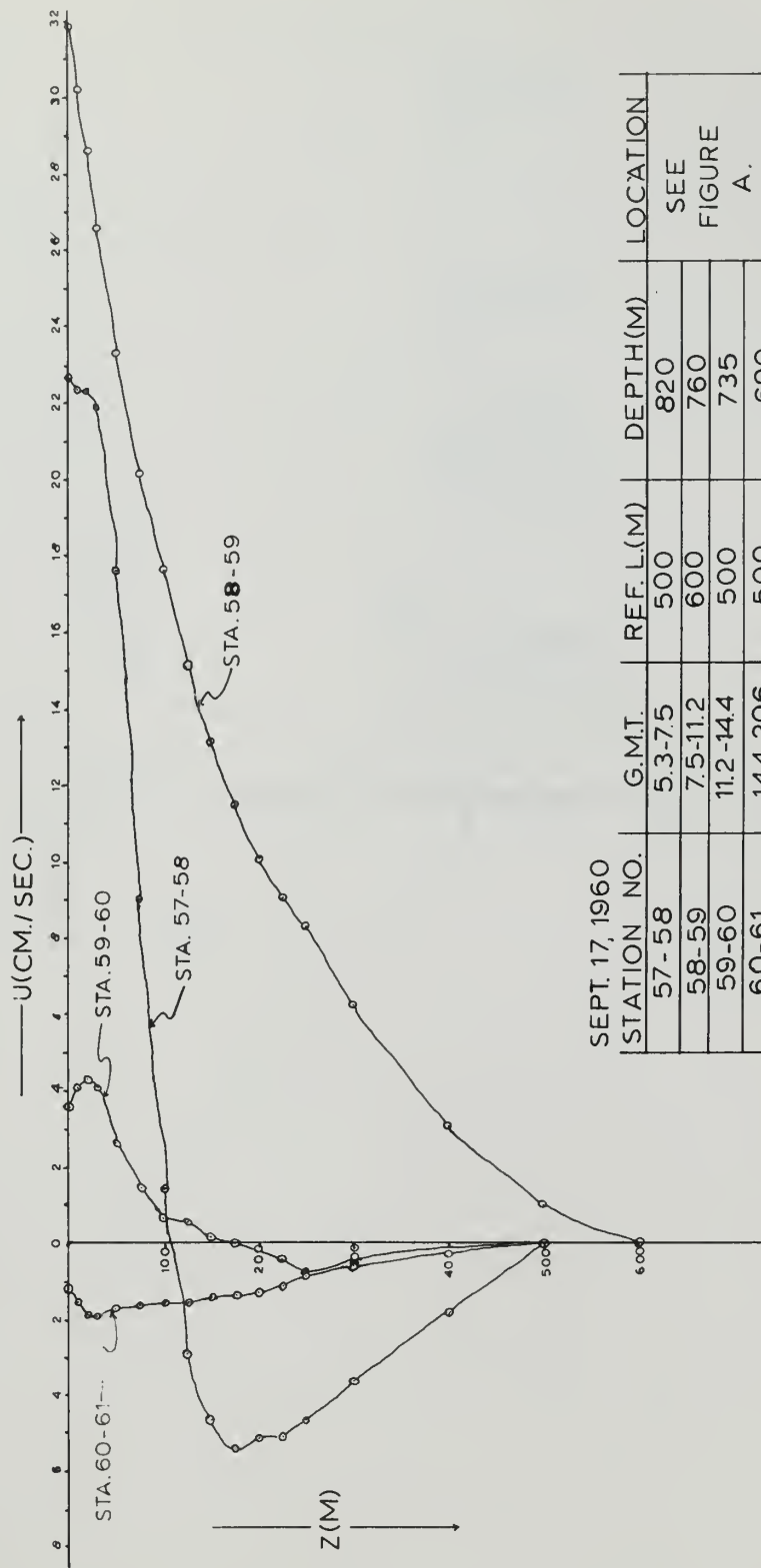


FIGURE B

Spring Data: 1 to 11 June 1963
 Location of Ocean Stations 24 through 27,
 and 41 through 51. Direction of surface
 current between stations as indicated

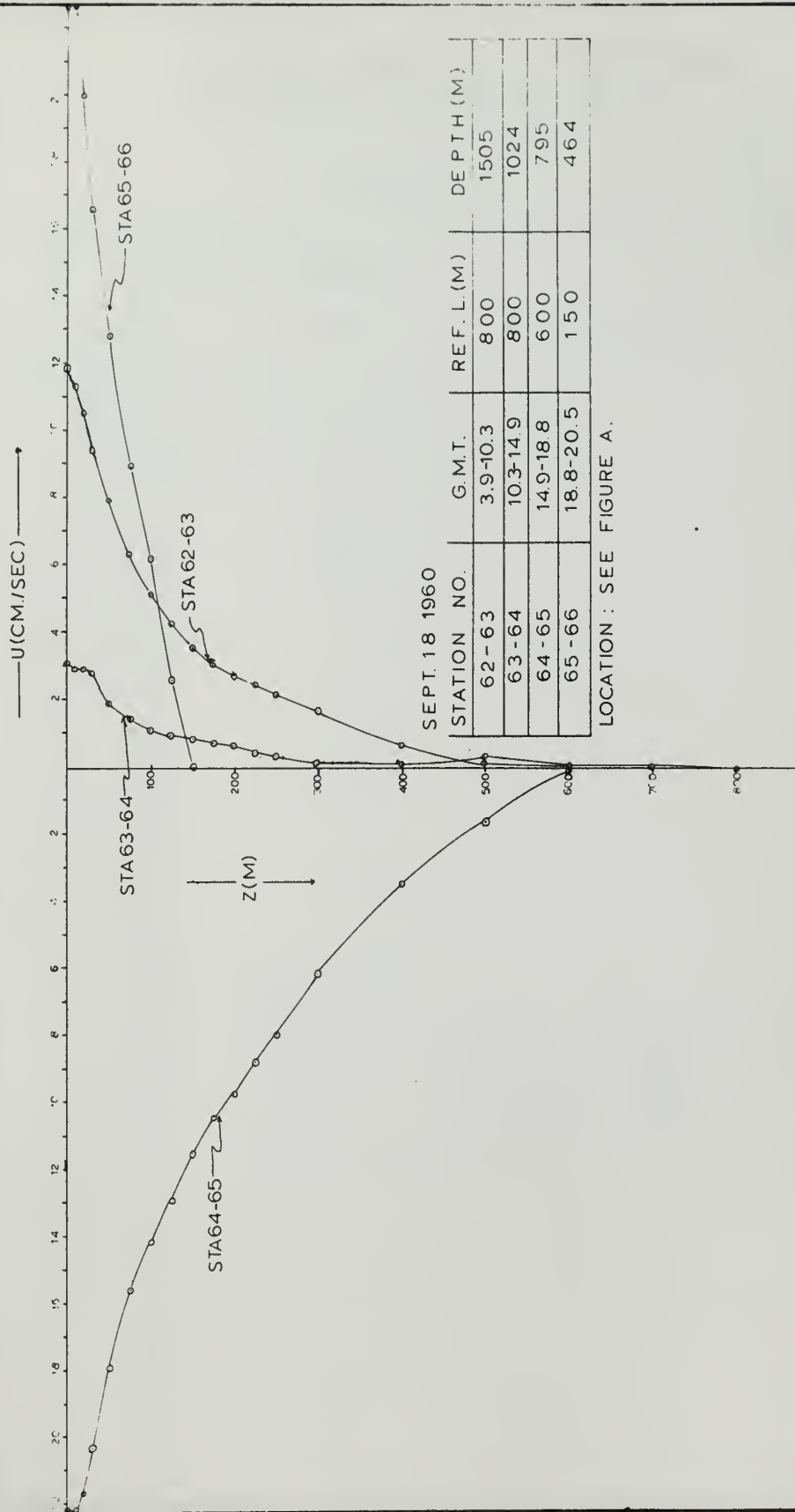
VELOCITY PROFILES AT STATION 57-58, 58-59, 59-60, 60-61



SEPT. 17, 1960

STATION NO.	G.M.T.	REF. L.(M)	DEPTH(M)	LOCATION
57-58	5.3-7.5	500	820	SEE FIGURE A.
58-59	7.5-11.2	600	760	
59-60	11.2-14.4	500	735	
60-61	14.4-20.6	500	690	

VELOCITY PROFILES AT STATION 62-63, 63-64, 64-65, 65-66

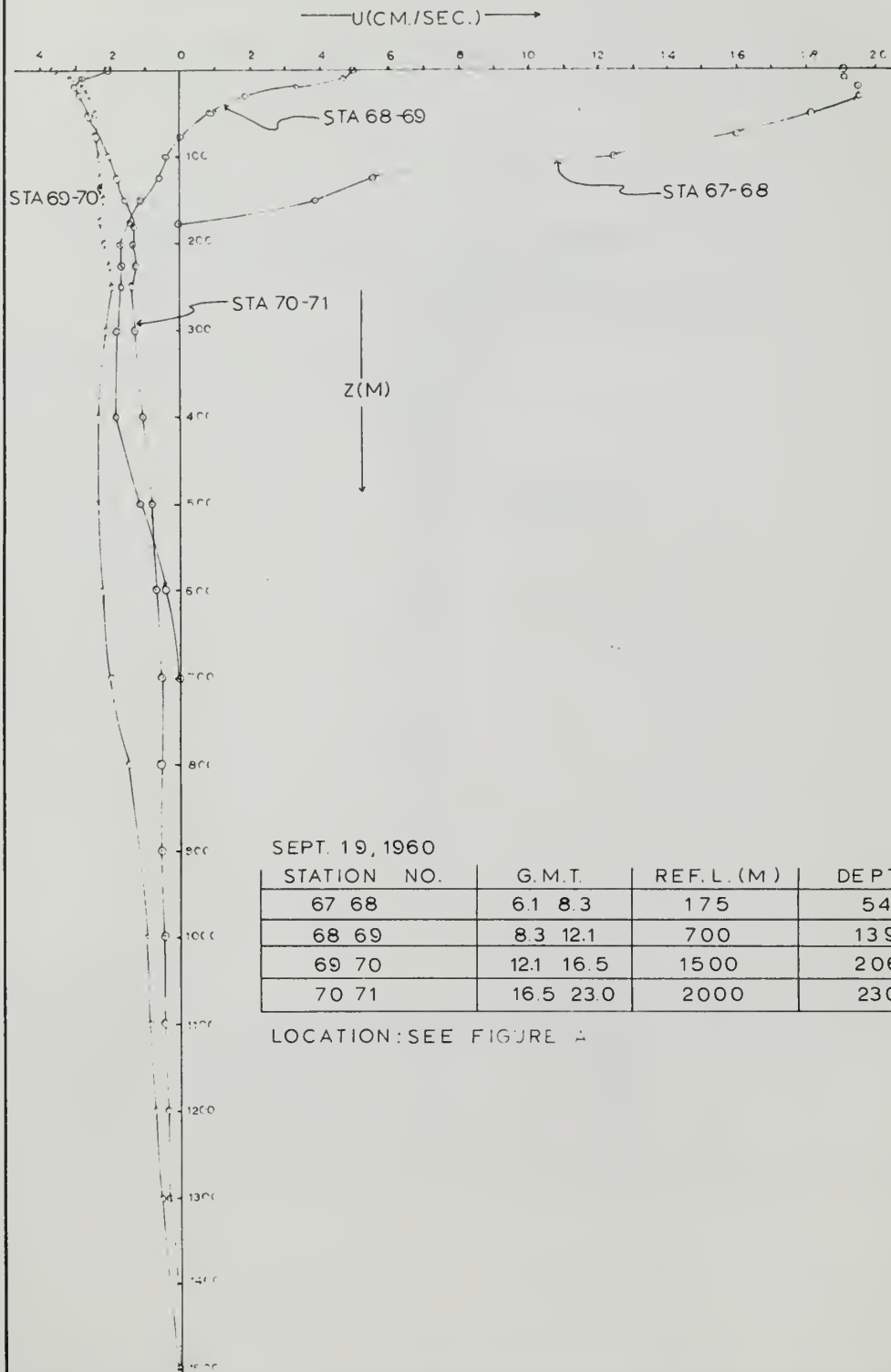


SEPT. 18 1960

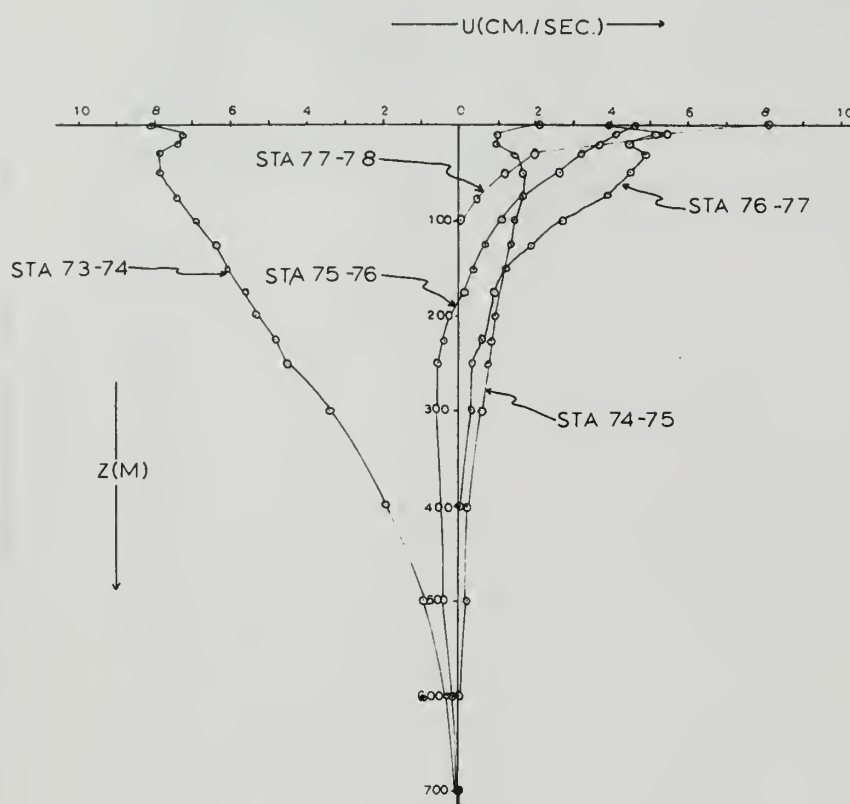
STATION NO.	G.M.T.	REF. L. (M)	DEPTH (M)
62-63	3.9-10.3	800	1505
63-64	10.3-14.9	800	1024
64-65	14.9-18.8	600	795
65-66	18.8-20.5	150	464

LOCATION : SEE FIGURE A.

VELOCITY PROFILES AT STATION 67-68, 68-69, 69-70, 70-71



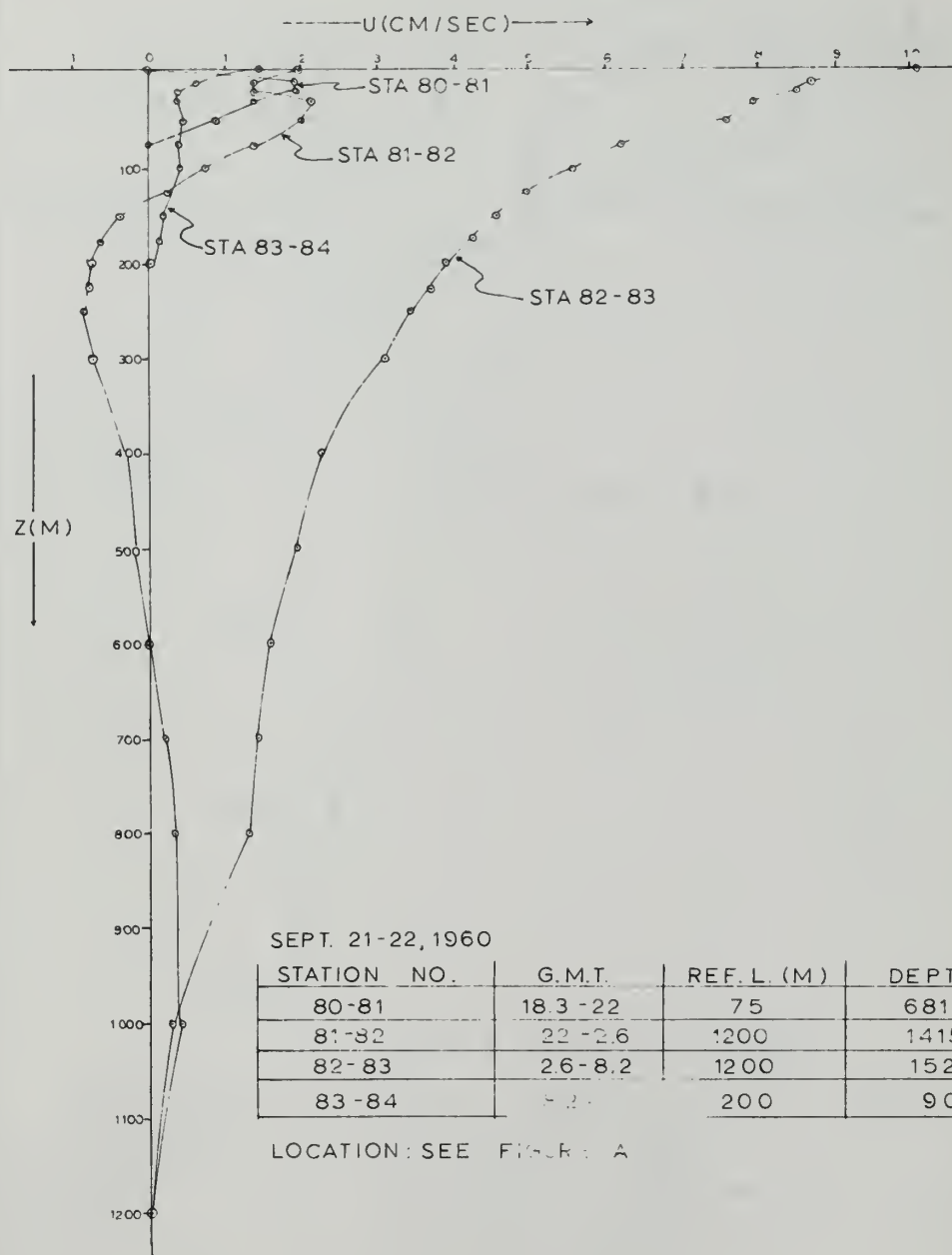
VELOCITY PROFILES AT STATION 73-74, 74-75, 75-76, 76-77, 77-78



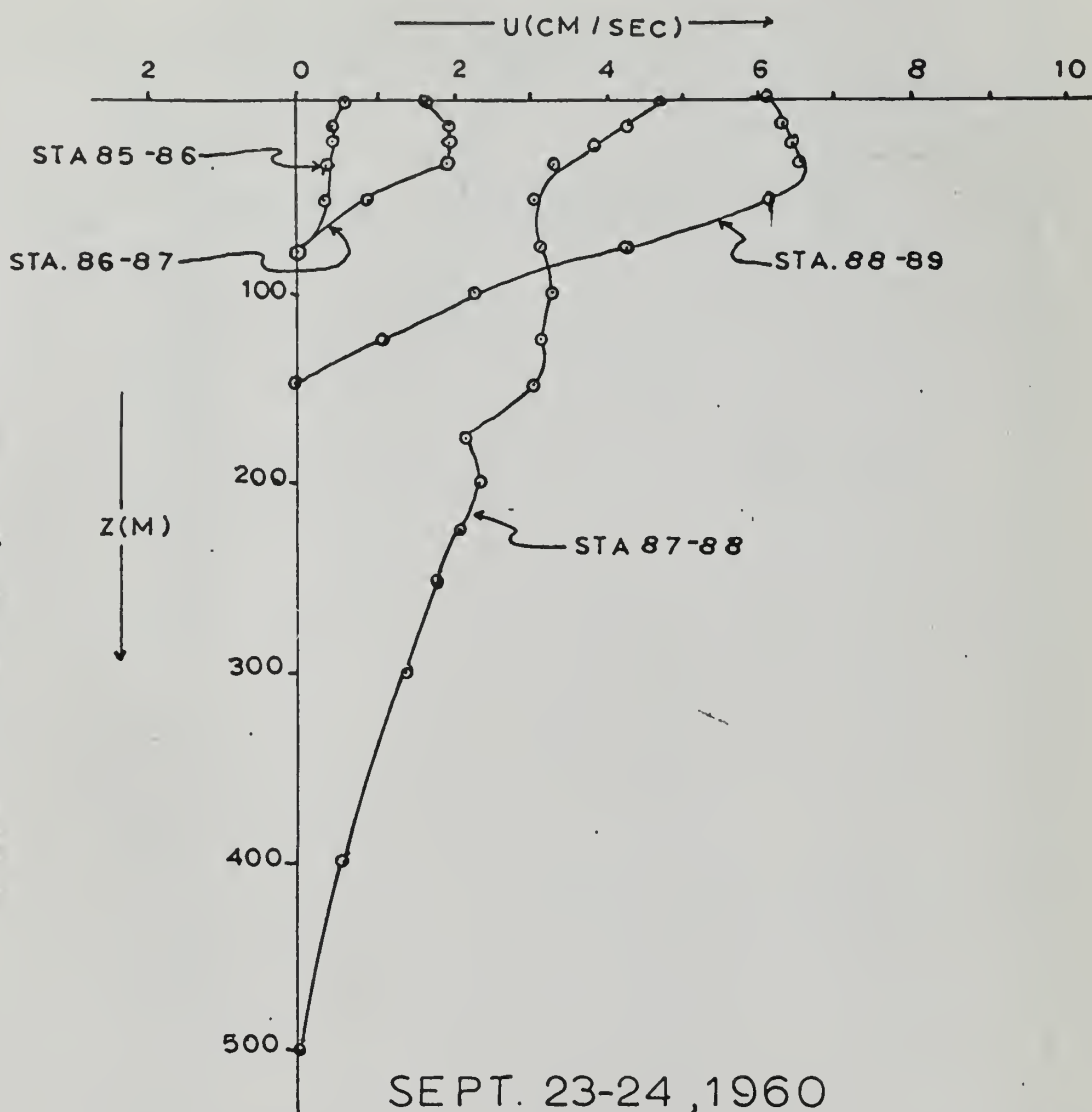
SEPT. 20-21, 1960

STATION NO.	G.M.T.	REF. L. (M)	DEPTH (M)	LOCATION
73-74	11.8-15.6	700	1504	SEE FIGURE A.
74-75	15.6-19.3	700	2080	
75-76	19.3-23.6	700	1506	
76-77	23.6-2.0	500	777	
77-78	2.0-4.3	100	350	

VELOCITY PROFILES AT STATION 80-81, 81-82, 82-83, 83-84



VELOCITY PROFILES AT STATION 85-86, 86-87, 87-88 , 88-89

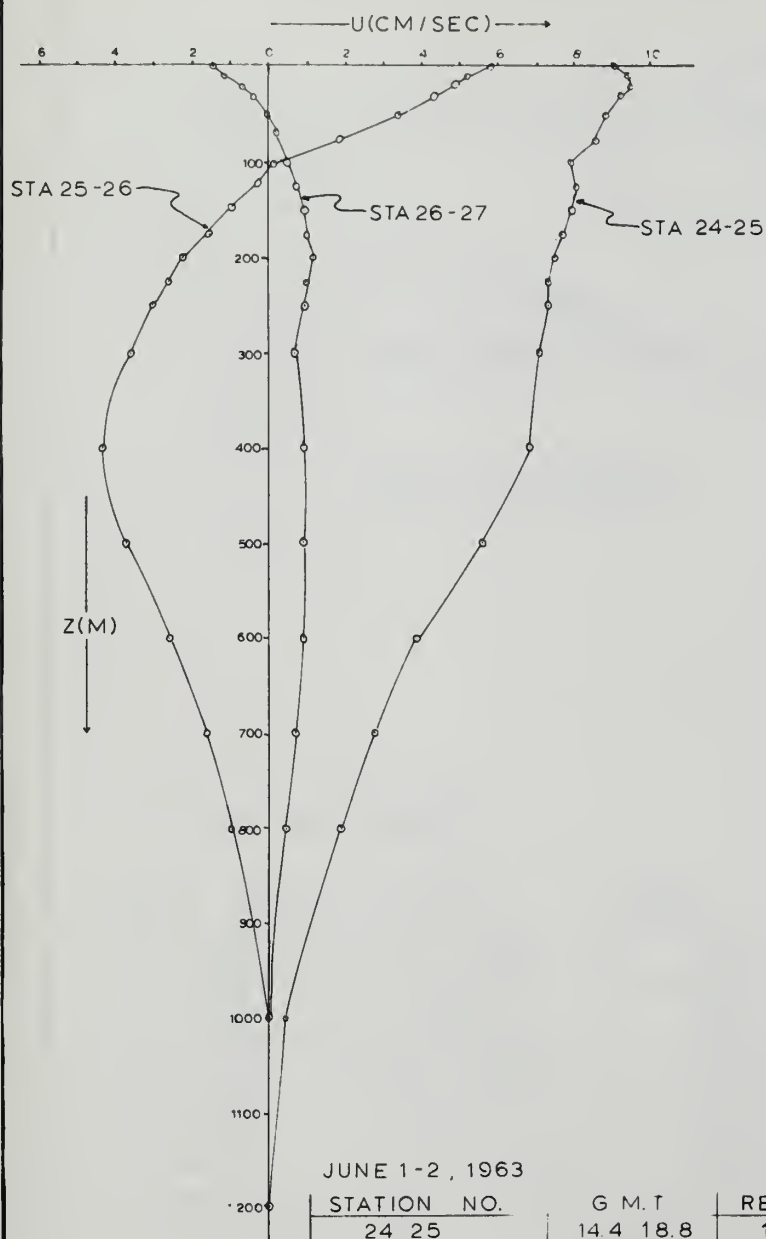


SEPT. 23-24, 1960

STA. NO.	NO	G.M.T.	REF. L.(M)	DEPTH(M)
85-86		5.8-9.9	75	139
86-87		9.9-15.7	125	389
87-88		15.7-20.8	500	576
88-89		20.8-0.2	150	361

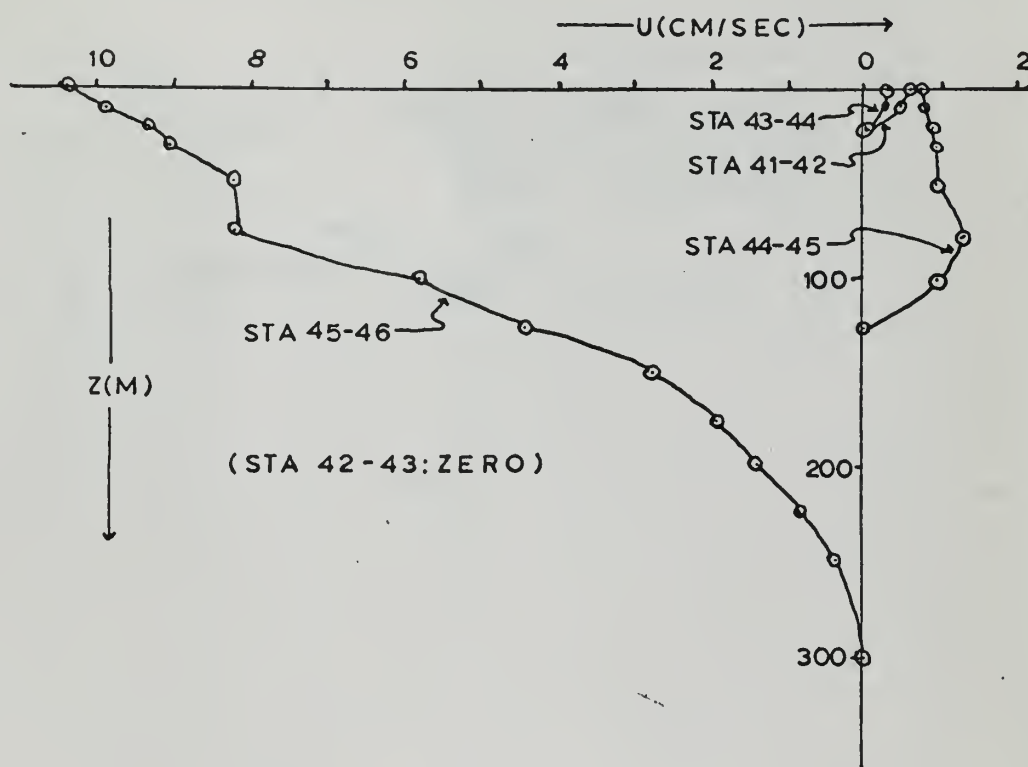
LOCATION: SEE FIGURE A

VELOCITY PROFILES AT STATION 24-25, 25-26, 26-27



LOCATION : SEE FIGURE B

VELOCITY PROFILES AT STATION 41-42, 42-43, 43-44, 44-45, 45-46

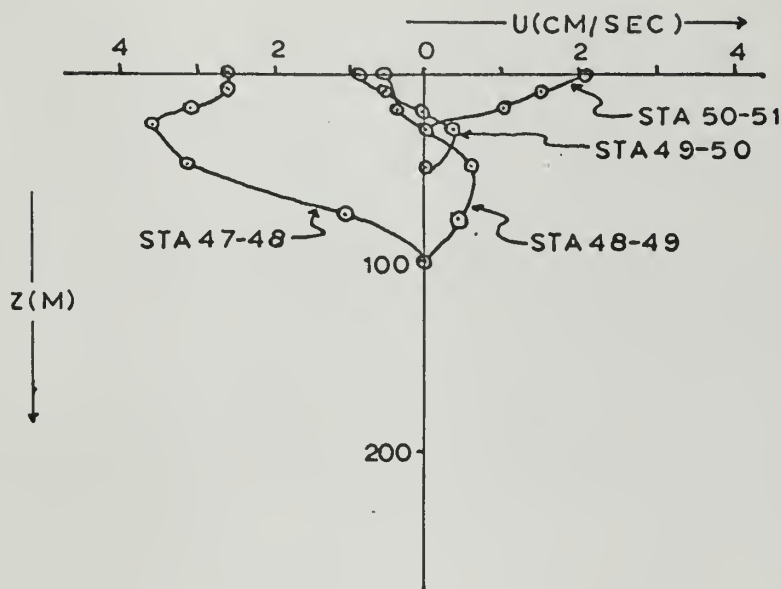


JUNE 10, 1963

STA NO.	G. M. T.	REF. L. (M)	DEPTH (M)
41-42	1.6-3.5	30	49
42-43	3.5-6.2	30	79
43-44	6.2-8.5	75	133
44-45	8.5-10.7	125	253
45-46	10.7-14.8	300	514

LOCATION: SEE FIGURE B

VELOCITY PROFILES AT STATION 47-48 , 48-49 , 49-50 , 50-51



JUNE 11 1963

STA. NO.	G. M. T.	REF. L.(M)	DEPTH(M)
47-48	3.3-5.2	175	223
48-49	5.2-8.2	100	149
49-50	8.2-10.5	50	88
50-51	10.5-12.6	30	52

LOCATION: SEE FIGURE B

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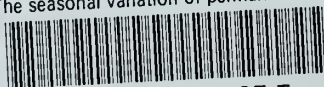
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